Hindcasting to measure ice sheet model sensitivity to initial states: Supplementary Information

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1 Climate Forcing

The hydrostatic atmospheric regional climate model HIRHAM5 (Christensen et al., 2006) provides monthly mean climatic mass balance and 2-m air temperature. HIRHAM is based on HIRLAM7 dynamics (Eerola, 2006) and ECHAM5 physics (Roeckner et al., 2003). HIRHAM5 is forced at the lateral boundaries using the European Centre for Medium-Range Weather Forecasts ERA-Interim reanalysis product (Dee et al., 2011) for the period 1989–2011. The regional climate model dynamically downscales the temperature and precipitation fields and computes the climatic mass balance at a resolution of 0.05° ($\sim$5.55 km) with 31 vertical levels and a time-step of 120 s in the dynamical scheme. HIRHAM5 has been validated both with ice core data and automatic weather station data and is shown to perform well over Greenland (Dethloff et al., 2002; Kiilsholm et al., 2004; Lucas-Picher et al., 2012).

2 Ice Sheet Model

Simulations are performed with the open-source Parallel Ice Sheet Model (PISM, www.pism-docs.org)[1], which is thermomechanically-coupled, polythermal, and includes a hybrid stress balance model (Bueler and Brown, 2009; Aschwanden et al., 2012). PISM has been used in a number of studies of the ice sheets on Greenland and Antarctica (e.g., Martin et al., 2011; Solgaard et al., 2011; Solgaard and Langen, 2012; Golledge et al., 2012).

At the ocean boundary, ice is calved-off at the initial calving position, which is held fixed throughout the simulations. Basal topography and ice thickness are from a new compilation by J. A. Griggs and J. L. Bamber, University of Bristol for the EU Ice2Sea project. At the basal boundary, geothermal flux varies in space (Shapiro and Ritzwoller, 2004), and a nearly-plastic power law (Schoof and Hindmarsh, 2010) relates bed-parallel shear stress, $\boldsymbol{\tau}_b$, and the sliding velocity, $\mathbf{u}_b$:

$$\boldsymbol{\tau}_b = -\tau_c \frac{\mathbf{u}_b}{|\mathbf{u}_b|^{(1-q)} u_0^q}, \tag{2.1}$$

where τ_c is the yield stress, q is the pseudo-plasticity exponent, and $u_0 = 100\,\mathrm{m\,a^{-1}}$ is a threshold speed. The basal material (till) is partially water saturated. Saturation, yield stress and modeled liquid water pressure within the till are related by the pore water pressure, p_w, given as a fraction of the overburden pressure:

$$p_w = \alpha\, w \rho g H \qquad \text{in} \qquad \tau_c = (\tan\phi)(\rho g H - p_w). \tag{2.2}$$

Here, H is the ice thickness, $\rho g H$ is the overburden pressure, and ϕ is the till friction angle. The relative amount of stored water in the till, w, comes from time-integrating the basal melt rate. Excess water drains when the thickness of stored water reaches 2 m. The coefficient α is the allowed pore water pressure fraction. The yield stress τ_c is also a function of the till friction angle ϕ, prescribed as a continuous function of the bed elevation, with $\phi = 5°$ for bed elevations lower than 300 m below sea level, $\phi = 20°$ for bed locations higher than 700 m above sea level, and changing linearly in between.

The enhancement factor, E, is a simple tuning parameter for ice dynamics. This factor is commonly used in ice sheet models to change the ice flow properties to account for the effect of the anisotropic nature of ice as well as impurities.

The three tuning parameters that are used to obtain a close fit to the observed surface speed are the enhancement factor E, the pseudo-plasticity exponent q, and the allowed pore water pressure fraction α. We use $(E, q, \alpha) = (3, 0.25, 0.95)$.

[1] All simulations used development revision `c52bdfd`

The three initialization procedures start from the same model state, obtained by running the model at a 20 km horizontal grid resolution with constant surface forcing for 50 ka while keeping the geometry fixed.

Horizontal and vertical model resolutions are 2 km and 10 m, respectively. The computational domain extends horizontally over 1500 km $\times$ 2800 km, and vertically over 4000 m and 2000 m in the ice and in the bedrock, respectively. For computational efficiency grid refinements are made during the initialization procedure. The runs are started on a 20 km grid, then refined from 20 km to 10 km at at -7'000 a, to 5 km at -2'000 a, to 2.5 km at -850 a, and finally to 2 km at -350 a. However, all the runs continue on their respective resolutions such that all initializations are available on all the listed grid resolutions. The run lengths are sufficiently long to remove unphysical transient noise resulting from the grid-refinement. Basal topography is remapped onto the respective grids using a first-order conservative algorithm. The correlation coefficient is consistent for all grid resolutions (Table 4.2). PISM uses an adaptive time-stepping scheme, and typical time-steps are about 2 days for a horizontal grid resolution of 2 km.

3 Flux-correction

The flux-correction method explained below is applied during the last 2,000 years of the flux-corrected paleo-climate initialization.

To obtain an ice sheet geometry in closer agreement with measurements we modify the climatic mass balance M. Let H_{tar} be this target thickness, and let H be the current model thickness. Recall that the mass continuity equation is

$$\frac{\partial H}{\partial t} = M - S - \nabla \cdot \mathbf{q}, \tag{3.1}$$

where M and S are the surface and basal mass balance, respectively, and $\nabla \cdot \mathbf{q}$ is the flux divergence. Replacing M with the modified surface mass balance yields the modified continuity equation

$$\frac{\partial H}{\partial t} = \tilde{M} - S - \nabla \cdot \mathbf{q}, \tag{3.2}$$

with

$$\tilde{M} = M + \Delta M = M + \beta \left(H_{\mathrm{tar}} - H \right), \tag{3.3}$$

where $\beta > 0$ is set to $0.05\,\mathrm{a}^{-1}$.

4 Supplementary Figures and Tables

Table 4.1: Consistency of the mass loss trend, b, with respect to grid resolution. Mean ($\bar{b}$), and standard deviation (b_σ) of the mass loss rates computed on 20, 10, 5, 2.5, and 2 km horizontal grid resolution.

	constant-climate	paleo-climate	flux-corrected
$\bar{b}$ [$\mathrm{Gt\,a^{-1}}$]	-140	-361	-210
b_σ [$\mathrm{Gt\,a^{-1}}$]	5	102	14

Table 4.2: Correlation coefficient, r, of the correlation between simulated and observed mass loss time series as a function of horizontal grid resolution, g.

g [km]		r [-]	
	constant-climate	paleo-climate	flux-corrected
20	0.910	0.984	0.882
10	0.883	0.977	0.882
5	0.918	0.984	0.988
2.5	0.920	0.986	0.989
2	0.919	0.986	0.988

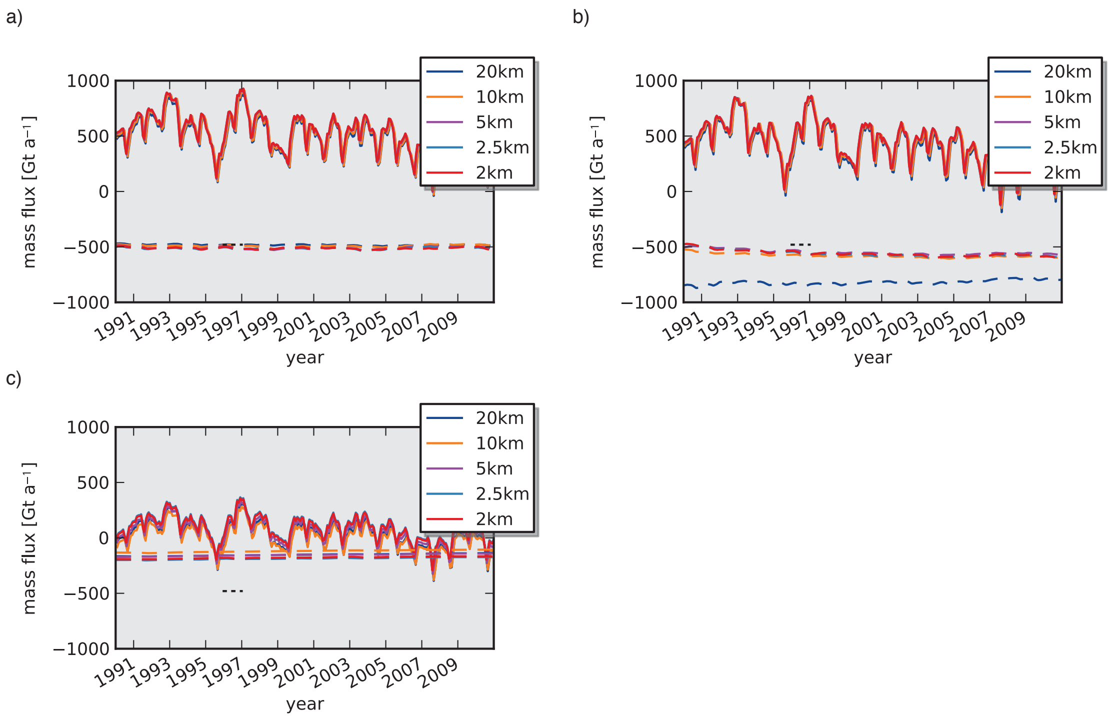

Figure 4.1: Simulated mass fluxes as a function of grid resolution from 1990 to 2008. a) constant-climate hindcast. b) paleo-climate hindcast. c) flux-corrected hindcast. Climatic mass balance (solid line), ice discharge (dashed line), and grounded basal melt (dashed-dotted line). For comparison the ice discharge estimate for 1996 (van den Broeke et al., 2009) is shown (black dotted line). Time-series are smoothed with a 13-month running-average filter.

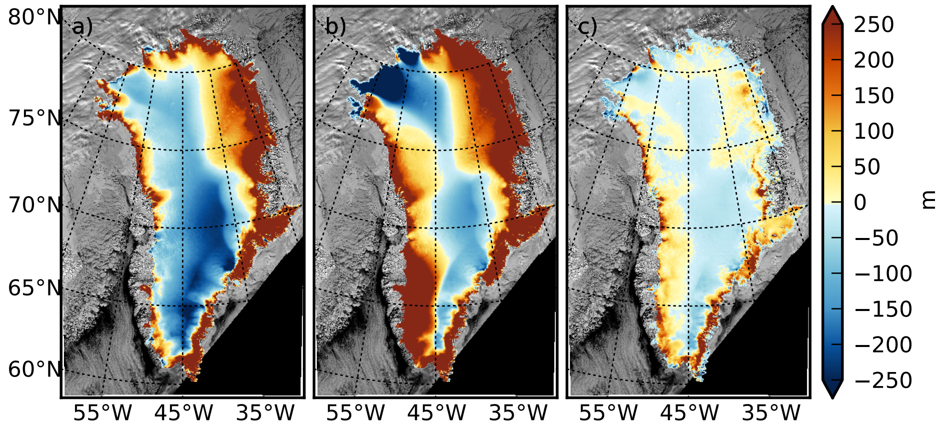

Figure 4.2: Absolute difference in ice thickness (model-observation). a) constant-climate initialization. b) paleo-climate initialization. c) flux-corrected paleo-climate initialization; ice thickness was used in the initialization, hence only shown for comparison. Blue and red colors indicate the model under- and over-estimates ice thickness, respectively. MODIS mosaic in the background is courtesy of M. Fahnestock.

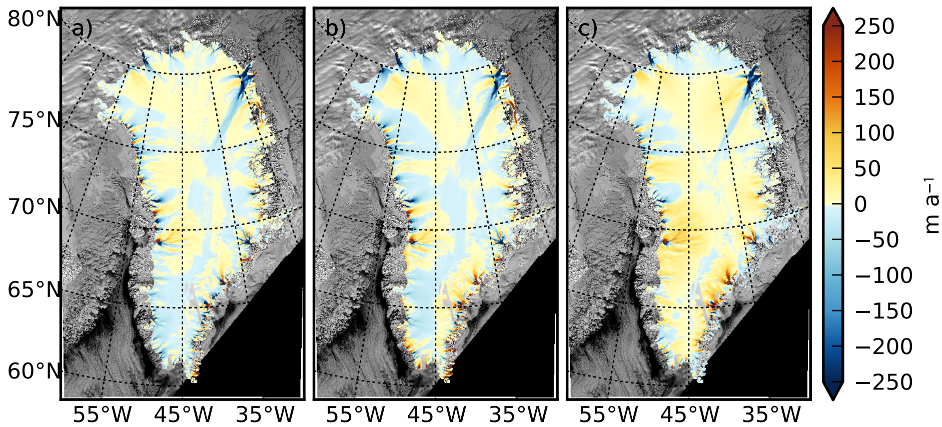

Figure 4.3: Absolute difference in surface speeds (model-observation). a, constant-climate initialization. b, paleo-climate initialization. c, flux-corrected paleo-climate initialization. Blue and red colors indicate the model under- and over-estimates surface speed, respectively. MODIS mosaic in the background is courtesy of M. Fahnestock.

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